The Jan Mayen Microplate Complex and the Wilson Cycle

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Abstract

The opening of the North Atlantic region was one of the most important geodynamic events that shaped the present-day passive margins of Europe, Greenland and North America. Although well-studied, much remains to be understood about the evolution of the North Atlantic, including the role of the Jan Mayen Microplate Complex (JMMC). Geophysical data provide an image of the crustal structure of this microplate and enable a detailed reconstruction of the rifting and spreading history. However, the mechanisms that cause separation of microplates between conjugate margins are still poorly understood. In this contribution, we assemble recent models of rifting and passive margin formation in the North Atlantic and discuss possible scenarios that may have led to formation of the JMMC. This event has likely been triggered by regional plate-tectonic reorganisations rejuvenating inherited structures. The axis of rifting and continental breakup and the width of the JMMC was controlled by old Caledonian fossil subduction/suture zones. Its length is related to E-W oriented deformation and fracture zones possibly linked to rheological heterogeneities inherited from pre-existing Precambrian terrane boundaries.

(end of abstract)

The North Atlantic region inspired some aspects of plate tectonic theory (Fig. 1). These include the Wilson Cycle which predicts the closure of oceans leading to continent-continent collision followed by their reopening along former sutures (Wilson 1966, Dewey & Spall 1975). The North Atlantic is often considered to be a text-book example of an ocean that opened along the former sutures of at least two temporarily distinct orogenic events – the Neoproterozoic Grenvillian-Sveconorwegian and the early Palaeozoic Caledonian-Variscan orogenies (Ryan & Dewey, 1997; Vauchez *et al.*, 1997; Bowling & Harry, 2001; Thomas, 2006; Misra, 2016). Nevertheless, some aspects of the North Atlantic geology remain enigmatic, such as the formation of the North Atlantic Igneous Province (NAIP) (Vink, 1984; White & McKenzie, 1989; Foulger & Anderson, 2005; Meyer *et al.*, 2007), the development of the volcanic passive margins (Franke, 2013; Geoffroy *et al.*, 2015), the formation of Iceland and the development of the Jan Mayen Microplate Complex (JMMC), also referred to as the Jan Mayen Microcontinent (Foulger *et al.*, 2003; Gaina *et al.*, 2009; Gernigon *et al.*, 2015). The JMMC comprises both oceanic and continental crust, probably highly thinned and magmatically modified (Kuvaas & Kodaira, 1997; Blischke *et al.*, 2016 and references therein). Large parts of it remain to be studied, however. Other continental fragments have been identified in the North Atlantic region (Nemčok *et al.*, 2016) and more may underlie parts of Iceland and/or the Iceland-Faroe Ridge (Fedorova *et al.*, 2005; Foulger, 2006; Paquette *et al.*, 2006; Gernigon *et al.*, 2012; Torsvik *et al.*, 2015).

Geological Setting of the North Atlantic region

Following the collision of Laurentia, Baltica and Avalonia in the Ordovician and Silurian (Roberts 2003, Gee *et al.* 2008, Leslie *et al.* 2008), and subsequent gravitational extensional collapse in the late orogenic phases (Dewey, 1988; Dunlap & Fossen, 1998; Rey *et al.*, 2001; Fossen, 2010), the North Atlantic region experienced lithospheric delamination and associated uplift over a period of 30-40 Ma, followed by a long period of rifting (Andersen *et al.*, 1991; Dewey *et al.*, 1993). Phases of extension and cooling transitioned into continental rifting that led to final continental breakup and seafloor spreading between Greenland and Europe in the early Palaeogene (Talwani & Eldholm 1977, Skogseid *et al.* 2000). During the late Mesozoic, continental breakup propagated simultaneously southward from the Eurasia Basin and northward from the Central Atlantic initially into the Labrador Sea- Baffin Bay rift system and then into the North Atlantic (Srivastava, 1978; Doré *et al.*, 2008). Whether rifting, continental breakup, and associated magmatism was initiated by active mantle upwelling, for example a deep mantle plume (White & McKenzie, 1989; Hill, 1991; Nielsen *et al.*, 2002; Rickers *et al.*, 2013) or plate-driven processes (Nielsen *et al.*, 2007; Ellis &

Stoker, 2014) (“bottom-up” or “top down” views) is still under debate (van Wijk *et al.*, 2001; Foulger *et al.*, 2005b; Lundin & Doré, 2005; Simon *et al.*, 2009; Peace *et al.*, 2017a).

The North Atlantic spreading axis initially comprised the Reykjanes Ridge, the Aegir Ridge, east of the JMMC and the Mohns Ridge farther north (Talwani & Eldholm, 1977; Nunns, 1982, Fig. 1). Independent rotation of the JMMC resulted in fan-shaped opening of the Norway Basin, during the Eocene (Nunns, 1982; Gaina *et al.*, 2009; Gernigon *et al.*, 2012). This reconfiguration led to a second phase of breakup and the separation of the JMMC from Greenland at approximately magnetic anomaly chron C7 (~24 Ma) (Vogt *et al.*, 1970; Gaina *et al.*, 2009; Gernigon *et al.*, 2015). After a period of simultaneous rifting on both the Aegir Ridge and the complex JMMC/proto-Kolbeinsey rift/ridge system (Doré *et al.*, 2008; Gaina *et al.*, 2009; Gernigon *et al.*, 2015), the Aegir Ridge was abandoned in the Oligocene and the spreading centre relocated to the west of the JMMC onto the Kolbeinsey Ridge. The present-day North Atlantic shows evidence for a dynamic contribution of the topography, requiring an anomalous pressure anomaly uplifting the lithosphere and possibly linked to the origin of Iceland (Schiffer & Nielsen, 2016).

Although the history of rifting in the North Atlantic is becoming increasingly better constrained, the mechanisms controlling the location, timing, and formation of rifts, fracture zones, and associated microcontinents are still poorly understood. The formation of the JMMC has been traditionally attributed to mantle plume impingement and subsequent lithospheric weakening (Müller *et al.* 2001). More recently it has been suggested to result from the breaching of lithosphere weakened as a result of pre-existing structures (*e.g.*, Schiffer *et al.* 2015b). The final separation of the JMMC is also spatially and temporally linked to enhanced magmatic activity and the subsequent formation of Iceland (Doré *et al.*, 2008; Tegner *et al.*, 2008; Larsen *et al.*, 2013; Schiffer *et al.*, 2015b) but it lacks the classic features of a volcanic passive margin (*e.g.*, underplating, seaward dipping reflectors) along its western continent-ocean boundary, conjugate to the East Greenland margin (Kodaira *et al.*, 1998; Breivik *et al.*, 2012; Peron-Pinvidic *et al.*, 2012; Blischke *et al.*, 2016). In this paper, we discuss the possible role of pre-existing structure and inheritance in formation of the JMMC as an extension to the Wilson Cycle and plate tectonic theory.

JAN MAYEN MICROPLATE COMPLEX

The JMMC has a bathymetric signature stretching over 500 km from north to south in the central part of the Norwegian-Greenland Sea (Fig. 1) (Gudlaugsson *et al.* 1988, Kuvaas & Kodaira 1997, Blischke *et al.* 2016). It is bordered to the north by the Jan Mayen Fracture Zone (JMFZ) and the volcanic complex of Jan Mayen Island. To the south, it is bordered by the NE coastal shelf of Iceland which is part of the Greenland-Iceland-Faroe Ridge (GIFR), a zone of shallow bathymetry approximately 1100 km length (Figs. 1 and 2). The JMMC separates the Norway Basin to the east from the Iceland Plateau to the west (Vogt *et al.* 1981, Kandilarov *et al.* 2012, Blischke *et al.* 2016).

The JMMC crust has been inferred to be continental primarily on the basis of seismic refraction data (Kodaira *et al.*, 1997; Kodaira *et al.*, 1998; Mjelde *et al.*, 2007a; Breivik *et al.*, 2012; Kandilarov *et al.*, 2012). However, for large areas of the JMMC crustal affinity remains uncertain, particularly near Iceland in the south (Breivik *et al.*, 2012; Brandsdóttir *et al.*, 2015) due to the lack of geophysical data and boreholes (see Gernigon *et al.*, 2015 and Blischke *et al.*, 2016 for data coverage). Fundamentally, the distribution of oceanic versus continental crust, as well as the nature of the deformation expected between the JMMC, Iceland and the Faroe continental block are unknown. Recent high-resolution aeromagnetic data and pre-rift reconstructions of the Norwegian-Greenland Sea show that the southern JMMC underwent extreme thinning during the first phase of breakup and, as it now has a width of ~250-300 km, 400% of extension has occurred compared to its pre-drift configuration (Gernigon *et al.* 2015). It seems unlikely that this extreme extension is entirely accommodated by the thinning of continental crust. We cannot rule out the possibility that the southern JMMC partly comprises igneous crust (Gernigon *et al.*, 2015) or exhumed mantle (Blischke *et al.*, 2016).

An oceanic fracture zone might be present south of the JMMC between the northeastern tip of the Iceland Plateau and the Faroe Islands in the southeast (i.e. the postulated Iceland-Faroe Fracture Zone, IFFZ, see Fig. 1 and 2, e.g. Blischke *et al.* 2016). However, an oceanic fracture zone or transform requires oceanic lithosphere on both sides and, given the uncertain crustal affinity this interpretation is speculative. A lineament exists north of the Iceland-Faroe Ridge (IFR. the part of the GIFR east of and including Iceland) but magnetic and gravity potential-field data do not provide conclusive evidence for a real oceanic transform or fracture zone (Fig. 3). Gernigon *et al.* (2012) showed that continuation of the magnetic chrons mapped in the Norway Basin and the high-magnetic trends observed along the IFR remain unclear, notably due to the low quality, the sparse distribution of the magnetic profiles along the IFR and later igneous overprint related to the formation of Iceland. No magnetic chrons are identified in the broad NE-SW magnetic lineations, especially west of the Faroe Platform. Additional magnetic disparities are associated with lateral variations of basement depth and possible discrete ridge jumps (e.g. Smallwood & White, 2002; Hjartarson et al., 2017). The GIFR comprises anomalous thick crust (>20-25 km) possibly associated with massive crustal underplating, which is generally attributed to increased magmatism (Staples et al., 1997; Richardson et al., 1998; Smallwood et al., 1999; Darbyshire et al., 2000; Greenhalgh & Kusznir, 2007). The origin and nature of the GIFR remains controversial (McBride et al., 2004), also because the crust shows atypical geophysical properties and differs from "normal" continental and oceanic crust (Bott, 1974; Foulger et al., 2003). A recent paper (Hjartarson et al., 2017) favours an oceanic origin of the IFR, but the authors do not exclude the presence of seaward dipping reflectors and old basement in the expected "oceanic domain". Some authors suggested that the excess thickness under Iceland may be partly attributed to buried continental crust possibly extending up to the JMMC and Iceland (Fedorova et al., 2005; Foulger, 2006). Continental zircons and geochemical analysis of lavas in southeast Iceland support the presence of continental material (Paquette et al., 2006; Torsvik et al., 2015). The Aegir Ridge and the Reykjanes Ridge might have never connected during the early stage of spreading of the Norway Basin involving complex overlapping spreading segments along the IFR. Such overlapping spreading ridges may have preserved continental lithosphere in between (Gaina et al., 2009; Gernigon et al., 2012, 2015; Ellis & Stoker, 2014). Ellis & Stoker (2014) suggested that no complete continental breakup along the IFR happened before the separation of the JMMC and the appearance of Iceland (first dated eruptions at ~18 Ma). Gernigon et al. (2015) suggested earlier breakup possibly between C22/C21 (~47 Ma) and C6 (~24Ma) during the onset of significant rifting in the southern part of the JMMC. The continental lithosphere east of Iceland (the IFR, Fig. 1) probably didn't entirely breach in the early rifting of the North Atlantic (e.g. C24r-C22, Early Eocene). To avoid further ambiguity, we refer to it as the Iceland-Faroe accommodation zone (IFAZ). Consequently, the IFAZ may characterize local continental transform margin segments, a diffuse strike-slip fault zone and/or a more complex oblique/transtensional continental rift system that initially formed along the trend pf the proto IFR.

MICROPLATE FORMATION

An aspect of the Wilson Cycle that requires more clarification (Thomas, 2006; Huerta & Harry, 2012; Buiter & Torsvik, 2014) is whether the locations of major, pre-existing structures can explain the formation, location and structure of microplates such as the JMMC (Schiffer *et al.* 2015a). Understanding the formation of continental fragments is crucial to understanding continental breakup (Lavier & Manatschal, 2006; Peron-Pinvidic & Manatschal, 2010). Microcontinents and continental ribbons represent one category of continental fragments produced during rifting and breakup (Lister *et al.*, 1986; Peron-Pinvidic & Manatschal, 2010; Tetreault & Buiter, 2014).

We follow the original definition of a microcontinent Scrutton (1976) that it must contain: (i) pre-rift basement rocks, (ii) crust and lithosphere of continental affinity, horizontally displaced from the original continent and surrounded by oceanic crust, and (iii) a distinct morphological feature in the surrounding oceanic basins. Such a system between two pairs of conjugate margins may also include isolated fragments of oceanic crust and lithosphere that deformed together before final and definitive isolation from the conjugate continents. To make a distinction, we call such a feature a microplate complex, and it can involve several sub-plates of oceanic and/or continental affinity. A true microcontinent will, therefore, comprise just one kind of microplate complex. The most important aspect of the present study is that such a microplate complex, like a true microcontinent, is separated from the main continental conjugate margins by two or more spreading ridges. The cause, history and processes leading to relocalisation of the complex are not well understood. Suggested mechanisms include the impact of a mantle plume (Müller *et al.*, 2001; Gaina *et al.*, 2003; Mittelstaedt *et al.*, 2008), global plate-tectonic reorganisation (Collier *et al.*, 2008; Gaina *et al.*, 2009), and ridge "jumps" that exploit inhomogeneities, weaknesses and rheological contrasts in the continental lithosphere after the abandonment of a previous spreading ridge (Abera *et al.* 2016, Sinha *et al.* 2016). This could be nascent or inherited underplating (Yamasaki & Gernigon 2010) and/or fossil suture zones Strike-slip mechanisms under different transtensional and transpressional stress regimes have also been proposed to generate microcontinents (Nemčok *et al.* 2016). Microplates can also result from crustal fragmentation during volcanic margin formation by large-scale continent-vergent faults formed/activated by strengthening of the deep continental crust – the so-called "C-Block" mechanism (Geoffroy *et al.* 2015).

Whittaker *et al.* (2016) proposed a model for microcontinent formation between Australia and Greater India whereby changes in plate motion direction caused transpression and stress buildup across large-offset fracture zones, leading to transfer of deformation to a less resistive locus (Fig. 4). Their proposed model is as follows. Initially NW-SE spreading separated Australia from Greater India with transtensional or strike-slip motion along the Wallaby-Zenith Fracture Zone from 133 Ma. A plume (Kerguelen) is postulated to have been in the vicinity and may have maintained and/or enhanced crustal weakening of the SE Greater India rifted margin. Reorganisations of motion between Australia and Greater India to a NNW-SSE direction at 105 Ma resulted in transpression along the NW-SE-oriented Wallaby-Zenith Fracture Zone. As a result, the spreading centre relocated to the west along the continental margin of India, calving off the Batavia and Gulden Draak microcontinents, and resulting in abandonment of the Dirck Hartog spreading ridge to the south (Fig. 4).

NORTH ATLANTIC – STRUCTURE AND INHERITANCE

The classic Wilson Cycle model envisages closure and reopening of oceans along continental sutures. In this model, breakup is thus guided by lithospheric inheritance from previous orogenesis (Wilson 1966, Dewey & Spall 1975). Inheritance, rejuvenation and control of pre-existing structure on localising deformation occurs on various scales and styles beyond large-scale breakup of continents (Holdsworth *et al.*, 1997; Manatschal *et al.*, 2015; Peace *et al.*, 2017b). Inherited features may include crustal or lithospheric thickness variations, structural and compositional heterogeneity across terrane boundaries, accreted terranes, sedimentary basins and/or intruded, metamorphosed and metasomatised material and fabrics. These heterogeneities may also cause thermal and rheological anomalies that vary in size, depth and degree of anisotropy, that can potentially be rejuvenated given the appropriate stresses (Krabbendam & Barr, 2000; Tommasi *et al.*, 2009; Manatschal *et al.*, 2015; Tommasi & Vauchez, 2015). Inheritance is an important control on rifting, passive-margin end-member style (*e.g.*, volcanic or non-volcanic) (Vauchez *et al.*, 1997; Bowling & Harry, 2001; Chenin *et al.*, 2015; Manatschal *et al.*, 2015; Schiffer *et al.*, 2015b; Svartman Dias *et al.*, 2015; Duretz *et al.*, 2016; Petersen & Schiffer, 2016), the formation of fracture zones, transform faults, transform margins (Thomas, 2006; Gerya, 2012; Doré *et al.*, 2015), magmatism (Hansen *et al.* 2009, Whalen *et al.* 2015), compressional deformation (Sutherland *et al.* 2000, Gorczyk & Vogt 2015, Heron *et al.* 2016), the breakup of supercontinents and supercontinent cycles (Vauchez *et al.*, 1997; Audet & Bürgmann, 2011; Frizon de Lamotte *et al.*, 2015).

Precambrian orogenies

In Canada, Greenland and Northwest Europe, multiple suturing events have built continental lithosphere that comprises Archean-to-early Proterozoic cratons surrounded by younger terranes. Preserved sutures and subduction zones in the interior of the cratons have survived subsequent amalgamation demonstrating that crustal and upper mantle heterogeneities may persist for billions of years (Balling 2000, van der Velden & Cook 2005). Terrane boundaries of any age may act as rheological boundaries that influence or control crustal deformation long after their formation and independently of subsequent plate motions. Major Precambrian terrane boundaries in the North Atlantic region are shown in Figure 2.

Multiple Precambrian suturing events have contributed to the amalgamation of the Baltic Shield in Scandinavia. The Lapland-Kola mobile belt formed by accretion of various Archean to Palaeoproterozoic terranes, including the oldest Karelian terrane (Gorbatschev & Bogdanova 1993, Bergh *et al.* 2012, Balling 2013). This was followed by the late Palaeoproterozoic Svecofennian accretion, the formation of the Transscandinavian Igneous Belt, and finally the Meso-Neoproterozoic Sveconorwegian orogeny (Gorbatschev & Bogdanova, 1993; Bingen *et al.*, 2008; Bergh *et al.*, 2012; Balling, 2013; Slagstad *et al.*, 2017).

Precambrian terranes are also preserved in Greenland, the oldest of which are Archean in age and include the North Atlantic and Rae Cratons (St-Onge *et al.* 2009). The components that together constitute the North Atlantic Craton formed 3850 – 2550 Ma (Polat *et al.* 2014) and the Rae Craton formed 2730 – 2900 Ma (St. Onge *et al.* 2009). Paleoproterozoic terranes in Greenland surround the North Atlantic Craton and include (i) the Nagssugtoqidian Orogen (Van Gool *et al.* 2002), (ii) the Rinkian Orogen (Grocott & McCaffrey 2016) and (iii) the Ketilidian Mobile Belt (Garde *et al.* 2002).

The Precambrian terranes of northeast Canada, Greenland and Scandinavia are thought to have formed as coherent mobile belts (Kerr *et al.*, 1996; Wardle *et al.*, 2002; St-Onge *et al.*, 2009). As Greenland and North America have not undergone significant relative lateral motions or rotation the interpretation of conjugate margins is relatively simple (Kerr *et al.*, 1996; Peace *et al.*, 2016). In contrast, whether or not Baltica has experienced rotation (Gorbatschev & Bogdanova 1993, Bergh *et al.* 2012) is currently unresolved.

Caledonian Orogeny

Formation of the Ordovician to Devonian Caledonian-Appalachian Orogen preceded rifting, ocean spreading and subsequent passive margin formation of the present-day North Atlantic. This Himalaya-style orogen involved at least two phases of subduction: (i) the early eastward-dipping Grampian-Taconian event and (ii) the late westward-dipping Scandian event that led to the assembly of part of Pangaea (Roberts 2003, Gee *et al.* 2008). During orogenesis the structural fabric of the crust and lithospheric mantle can be reoriented resulting in fabric anisotropy that localises subsequent deformation (Tommasi *et al.*, 2009; Tommasi & Vauchez, 2015).

High-velocity, lower-crustal bodies (HVLCB) are observed along many passive continental margins (Lundin & Doré, 2011; Funck *et al.*, 2016a) and have been traditionally associated with magmatic underplating or intrusions into the lower crust of passive margins during breakup (Olafsson *et al.* 1992, Eldholm & Grue 1994, R. Mjelde *et al.* 2007, White *et al.* 2008, Thybo & Artemieva 2013). However, with improved data alternative interpretations have been proposed such as syn-rift serpentinisation of the uppermost mantle under passive margins (Ren *et al.*, 1998; Reynisson *et al.*, 2010; Lundin & Doré, 2011; Peron-Pinvidic *et al.*, 2013). It has also been suggested that part of the continental HVLCB may be remnants of inherited metamorphosed crust or hydrated meta-peridotite that existed prior to initial rifting and continental breakup (Gernigon *et al.*, 2004; Gernigon *et al.*, 2006; Fichler *et al.*, 2011; Wangen *et al.*, 2011; Mjelde *et al.*, 2013; Nirrengarten *et al.*, 2014).

Mjelde *et al.* (2013) have identified a number of such “orogenic” HVLCB along different parts of the North Atlantic passive margins (the South- and Mid-Norwegian margin, East Greenland margin, SW Barents Sea margin, Labrador margin), which may have higher than normal upper mantle velocities (Vp > 8.2 km/s). These may comprise eclogitised crust and be part of the Iapetus Suture. Petersen & Schiffer (2016) proposed a mechanism to explain the presence of old inherited HVLCB beneath the rifted margins and concluded that they could represent preserved and subsequently deformed pre-existing subduction/suture zones that were activated during rifting and continental breakup. Eclogite in a fossil slab has a similar but weaker rheology than the surrounding "dry olivine" lithosphere (after Zhang & Green, 2007), while a fossil, hydrated mantle wedge acts as an effective and dominant weak zone. Eclogites of the Bergen Arcs (Norway) show softening due to fluid infiltration Jolivet *et al.* (2005). These ultra-high velocity HVLCB (ultra-HVLCB) are distributed primarily along the mid-Norwegian margin and the Scoresbysund area in East Greenland (Mjelde *et al.*, 2013). This suggests that at least one fossil subduction zone may have been subject to rift-related deformation and exhumation (Petersen & Schiffer 2016).

Structures in the Central Fjord area of East Greenland (Schiffer *et al.* 2014), the Flannan reflector in northern Scotland (Snyder & Flack 1990, Warner *et al.* 1996) and the Danish North Sea (Abramovitz & Thybo 2000) have been interpreted as preserved orogenic structures of Caledonian age (i.e. fossil subduction or suture zones) (Fig. 2). Schiffer *et al.* (2015a) proposed that the Central Fjord structure and the Flannan reflector once formed a contiguous eastward-dipping subduction zone, possibly of Caledonian age, that may have influenced rift, magmatic, and passive-margin evolution in the North Atlantic (Figure 2). Combined geophysical-petrological modelling of the Central Fjord structure suggests it comprises a relict hydrated mantle wedge associated with a fossil subduction zone (Schiffer *et al.* 2015b, Schiffer *et al.* 2016). The most recent Caledonian subduction event was associated with the Scandian phase leading to the westward subduction of Iapetus crust (Roberts 2003, Gee *et al.* 2008). Evidence of this subduction zone in the form of a preserved slab has not been detected in the lithospheric mantle of the Norwegian Caledonides. However, structures in the crust and upper mantle in the Danish North Sea detected by the Mona Lisa experiments (Abramovitz & Thybo 2000) might be the trace of this subduction. HVLC indicative of eclogite along the Mid-Norwegian margin (Mjelde *et al.*, 2013) and Norwegian North Sea (Christiansson *et al.*, 2000; Fichler *et al.*, 2011) might also represent deformed remnants of the Scandian subduction.

Fracture and accommodation zones

The JMMC is bound by two tectonic boundaries including the East and West Jan Mayen Fracture Zones in the north and the postulated Iceland-Faroe accommodation zone (IFAZ) in the south. These tectonic boundaries accommodated and allowed the non-rigid microplate to move independently from the surrounding North Atlantic oceanic domains (Gaina *et al.*, 2009; Gernigon *et al.*, 2012, 2015).

Relationships between pre-existing structures and the formation of large-scale shear and fracture zones, oceanic transforms or other accommodation/deformation zones have been proposed in previous work (Mohriak & Rosendahl, 2003; Thomas, 2006; Taylor *et al.*, 2009; de Castro *et al.*, 2012; Gerya, 2012; Bellahsen *et al.*, 2013; Gibson *et al.*, 2013). The location, orientation and nature of fracture zones in the North Atlantic may be linked to lithospheric inheritance (Behn & Lin, 2000). For example, the Charlie-Gibbs Fracture Zone between Newfoundland and the British/Irish shelf has been linked to the location of the Iapetus suture and inheritance of compositional and structural weaknesses (Tate 1992, Buiter & Torsvik 2014). The Bight Fracture Zone might be linked to the Grenvillian front, which is exposed in Labrador (Lorenz *et al.* 2012).

The IFAZ could represent a complex discontinuity zone along the present-day IFR. Along this transition zone between the Reykjanes, Aegir and Kolbeinsey ridges fragments of continental crust may be preserved together with discontinuous and/or overlapping oceanic fragments later affected by significant magmatic overprint (the Icelandic “swell”, Bott, 1988). In the geodynamic context, it may have formed along the fossil subduction zone proposed to have existed between the East Greenland and British/Irish margins (Fig. 2). It has also been proposed that it may have comprised part of the “Kangerlussuak Fjord tectonic lineament”, a NW-SE-oriented lineament in east Greenland (Tegner *et al.* 2008).

Other deformation zones may correlate with Precambrian basement terrane boundaries in Scandinavia. These are overprinted by Caledonian deformation, obscuring older relationships (cf. CDF in Fig. 2) and generating new orogenic fabrics (Vauchez *et al.*, 1998). The westward extrapolation of the northern Sveconorwegian suture may correlate with the East Jan Mayen Fracture Zone (EJMFZ), whilst extrapolation of the Svecofennian-Karelian suture may correspond to the formation of the Senja Fracture Zone (SFZ) (Doré *et al.* 1999, Fichler *et al.* 1999, Indrevær *et al.* 2013). Extrapolation of the Karelian-Lapland Kola terrane suture converges with the complex DeGeer Fracture Zone that marks the transition of the North Atlantic to the Arctic Ocean (Engen *et al.* 2008). These correlations suggest that Precambrian basement inheritance localises strain during initial continental rifting. However, the exact location and grade of deformation of Precambrian sutures under the Caledonides and the highly stretched continental margins is often poorly known or not known at all. Thus, any correlation is speculative and requires future work.

Iceland and magmatic evolution

Factors including the thermal state of the crust and mantle, small scale convection, upwelling, composition, volatile content, and lithospheric and crustal structure may all play roles (King & Anderson, 1998; Asimow & Langmuir, 2003; Korenaga, 2004; Foulger *et al.*, 2005a; Hansen *et al.*, 2009; Brown & Lesher, 2014; Chenin *et al.*, 2015; Hole & Millett, 2016).

Inheritance may influence the amount of volcanism produced in the North Atlantic because volcanic passive margins preferentially develop in regions of heterogeneous crust where Palaeozoic orogenic belts separate Precambrian terranes. Inversely, magma-poor margins often develop in the interiors of orogenic belts with either uniform-Precambrian or younger-Palaeozoic crust (Bowling & Harry, 2001). For example, the intersection of the East Greenland-Flannan fossil subduction zone with the North Atlantic rift axis correlates spatially and temporally with pre-breakup magmatism, the formation of JMMC and the occurrence of the Iceland melt anomaly along the sub-parallel GIR (Schiffer *et al.*, 2015b).

Prior to breakup (ca. 55 Ma), magma was dominantly emplaced along and south-west of the proposed East Greenland-Flannan fossil subduction zone (Fig. 2) (Ziegler, 1990; Torsvik *et al.*, 2002). This may be partly an effect of the south-to-north "unzipping" of the pre-North Atlantic lithosphere. Other processes that produce enhanced mantle melting are increased temperature, mantle composition and active asthenospheric upwelling (Brown & Lesher, 2014). The zonation of areas with and without magmatism may suggest that the proposed structure is a boundary zone between lithospheric blocks of different composition and rheology that react differently to applied stresses. Different relative strength in crust and mantle lithosphere, for instance, could cause depth dependent deformation, where thinning is focussed in the mantle lithosphere (Huismans & Beaumont 2011). Petersen & Schiffer (2016) demonstrated that extension of orogenic lithosphere with thickened crust (>45 km) leads to depth-dependent thinning where the mantle lithosphere breaks earlier than the crust and as a result encourages pre-breakup magmatism. Indirectly, sub-continental mantle heterogeneities may encourage localisation of deformation leading to rapid and sudden increase in lithospheric thinning (Yamasaki & Gernigon, 2010). These processes could contribute to pre-breakup adiabatic decompression melting (Petersen & Schiffer 2016). Enhanced magmatism could also be caused by a lowered solidus due to presence of eclogite (Foulger *et al.*, 2005a), water in the mantle (Asimow & Langmuir 2003) or CO_2 (Dasgupta & Hirschmann, 2006). Atypical magmatism is, surprisingly, observed along the interpolated axis of the proposed fossil subduction zone than elsewhere. It currently coincides with the GIFR where igneous crustal thickness is inferred to be greatest (Bott, 1983; Smallwood *et al.*, 1999; Holbrook *et al.*, 2001; Mjelde & Faleide, 2009; Funck *et al.*, 2016b). However, it is unclear whether the entire thickness of "Iceland type crust" (Bott, 1974; Foulger *et al.*, 2003) has crustal petrology (Foulger *et al.*, 2003; Foulger & Anderson, 2005).

Higher water contents have been recorded in basalts and volcanic glass in the vicinity of the fossil subduction zone (the Blosseville Kyst, East Greenland, Iceland and one sample from the Faroe Islands, see Fig. 2) than in regions further away from Iceland (West Greenland, Hold with Hope, Reykjanes Ridge) (Jamtveit *et al.* 2001, Nichols *et al.* 2002). This is consistent with a hydrated upper mantle source as a consequence of melting Caledonian subducted materials (Schiffer *et al.* 2015a). Water in the mantle may also contribute to enhanced melt production and thus unusually thick igneous crust (Asimow & Langmuir 2003).

The formation of the Iceland Plateau (>18 Ma) followed extinction of the Aegir Ridge and full spreading being taken up on the Kolbeinsey Ridge (Dore *et al.* 2008). This spreading ridge migration was contemporaneous with far-field plate tectonic reconfigurations, cessation of seafloor spreading in the Labrador-Baffin Bay system (Chalmers & Pulvertaft 2001) and a global change of Greenland plate motion from SW-NE to W-E (Gaina *et al.*, 2009; Abdelmalak *et al.*, 2012).

AN INHERITANCE MODEL FOR FORMATION OF THE JMMC

We propose a new tectonic model for formation of the JMMC that links rejuvenation of old and pre-existing orogenic structures to global plate tectonic reconfigurations. In our model a change in the orientation of the regional stress field in the Eocene rejuvenated pre-existing structures with favourable orientations. This caused relocalisation of extension and spreading ridges resulting in the formation of a microplate between the large European and American/Greenland continental plates. Our model closely follows that of Whittaker *et al.* (2016), with the extension that a fossil subduction zone is utilised as a physical and compositional weak zone that helps to accommodate a second axis of breakup (Fig. 5). Plate tectonic reorganisations and rejuvenation of pre-existing structures may not be the only controls on continental breakup, but they may be the dominant ones in the case of the JMMC. In areas where no microplate formation is observed continental breakup followed the youngest, weakest Caledonian collision zone, the Scandian, west-dipping subduction in Scandinavia. This may have been better aligned with the ambient stress field during rifting and/or breakup. Following the model of Petersen & Schiffer (2016), the remnants of this subduction zone or other inherited orogenic structures may now be distributed along the Mid-Norwegian margin as pre-breakup HVLCB (Christiansson *et al.*, 2000; Gernigon *et al.*, 2006; Fichler *et al.*, 2011; Wangen *et al.*, 2011; Mjelde *et al.*, 2013; Nirrengarten *et al.*, 2014; Mjelde *et al.*, 2016). The subduction zone was already deformed in the Norwegian North Sea by rifting subsequent to the Permo-Triassic and is still preserved as a large HVLCB beneath the North Sea rift (Christiansson *et al.* 2000, Fichler *et al.* 2011). A stronger, east-dipping subduction zone in East Greenland, may also have been deformed but did not accommodate breakup. Continental rifting and possible overlapping of the Reykjanes and Mohns ridge leading initiating the JMMC formation (Gernigon *et al.*, 2012, 2015) may have been promoted by the presence of this deep-rooted weak zone.

The Caledonian and Grenvillian orogenic fabric and major associated structures are generally parallel to the NNE-SSE trend of rifting in the North Atlantic with some exceptions, such as the opening of Labrador Sea. Older terrane boundaries are close to perpendicular. Young Caledonian structures define the axis of rifting and continental breakup. This can be explained by the presence of deep, weak eclogite-facies roots along the axis of the Caledonian Orogen, and extensional collapse of the Caledonian mountain range causing earlier extension to initiate perpendicular to the axis of collision (Ryan & Dewey, 1997; Rey *et al.*, 2001). Precambrian structures are still preserved in stable cratons surrounded by orogens and mobile belts. Once rifting occurs, lateral weaknesses and rheological boundaries control segmentation of the rift axis and eventually influence the formation of across-strike deformation zones of different kinds, *e.g.*, fracture and transform zones, diffuse/oblique/transtensional rift and ridge systems.

Our suggested scenario for the formation of the JMMC complements the established Wilson Cycle concept. We propose that reactivation and petrological variation of inherited structures of different ages, coupled with changes in the regional/global stress regime, controlled microplate formation in the following sequence of events (see also Fig. 6):

1. Early Palaeocene: Rifting propagates from the Central Atlantic into the Labrador Sea - Baffin Bay rift system (Roest & Srivastava, 1989; Chalmers & Pulvertaft, 2001; Peace *et al.*, 2016)
2. Early Eocene (Fig. 6b): Change in Labrador Sea-Baffin Bay spreading direction from NW-SE to W-E (Abdelmalak *et al.*, 2012) and onset of seafloor spreading in the northeast Atlantic (Gaina *et al.*, 2009). This was possibly related to the far-field stress field applied by the collision of Africa and Europe (Nielsen *et al.*, 2007) and/or to the relocation of the postulated Iceland plume (Skogseid *et al.*, 2000; Nielsen *et al.*, 2002).
3. The NW-SE stress field in the North Atlantic between Greenland and Scandinavia would have favoured deformation on deep structures associated with the Iapetus Suture on the Norwegian margin rather than the East Greenland margin with the proposed fossil subduction zone (Fig. 2). Thus, initial breakup is generally parallel to and in the vicinity of the Iapetus Suture.
4. The Iceland-Faroe Accommodation Zone (IFAZ) forms as the southern limit of the JMMC and may be linked to localisation of strain along the proposed fossil subduction zone or other potential rheological boundaries. No continental breakup occurred between Iceland and the Faroe Islands (Iceland Faroe Ridge), with underlying, uninterrupted but thinned, continental lithosphere (Ellis & Stoker, 2014).
5. Mid-late Eocene: Accellerated extension occurred in the southern part of the JMMC and local reorganisation of the Norway Basin spreading system (Gernigon *et al.* 2012, 2015) developed around 47 Ma (Fig. 6c) A first phase of magmatism between Greenland and the proto-JMMC was initiated (Tegner *et al.*, 2008; Larsen *et al.*, 2014). In the southern JMMC, isolated spreading cells possibly developed before steady state development of the Kolbeinsey Ridge.
6. Late Eocene - early Oligocene (Fig. 6c): A major plate tectonic reorganisation including a change from NW-SE to NE-SW plate motion coincident with abandonment of seafloor spreading along the Labrador Sea-Baffin Bay system

and consequent cessation of anti-clockwise rotation of Greenland (Mosar *et al.*, 2002; Gaina *et al.*, 2009; Oakey & Chalmers, 2012). This change in plate motion results in deformation along the fracture zones and transpression on the IFAZ.

7. Locking of the IFAZ triggered continental breakup between Greenland and the proto-JMMC subsequent to continental rifting between them. This is consistent with the microplate model of Whittaker *et al.* (2016) for the Indian Ocean. Rotational rifting between Greenland and the proto-JMMC started much earlier (c. 47-48 Ma) than abandonment of the Labrador Sea-Baffin Bay spreading system (c. 40 Ma) and breakup between Greenland and the JMMC (33-24 Ma).
8. Ultraslow spreading continued on the Aegir Ridge after ca. 31 Ma (Mosar *et al.*, 2002; Gaina *et al.*, 2009; Gernigon *et al.*, 2015), while drastic rifting and possible embryonic spreading developed south of the proto-JMMC until steady state spreading along Kolbeinsey Ridge was completely established at 24 Ma (Vogt *et al.*, 1970; Doré *et al.*, 2008; Gernigon *et al.*, 2012).
9. The Aegir Ridge was abandoned with all plate motion accommodated by the Kolbeinsey Ridge after 24 Ma, separating the proto-JMMC from East Greenland (Fig 6d). The West Jan Mayen Fracture Zone, the eastern branch of which had already been established during the opening of the Norway Basin, then connected the Kolbeinsey Ridge with the Mohns Ridge north of the JMMC.

SUMMARY

We propose a new model for formation of a microplate complex as an extension to the established Wilson Cycle concept. The new model invokes rejuvenation of major pre-existing structures by plate-driven processes controlling both breakup and JMMC formation.

The initial axis of continental breakup exploited lithospheric weaknesses associated with the Iapetus Suture (Fig. 6 a,b). These structures were particularly susceptible to deformation due to their preferential orientation with respect to the NW-SE to W-E oriented extensional stress field. Fracture zones and strike-slip/oblique zones of deformation delineate the later-forming JMMC. The IFAZ represents one of these zones and may have formed along an old subduction zone. The origin of the IFAZ remains poorly defined because of poor data coverage. However, it is likely that despite extreme thinning of the continental lithosphere no continental breakup occurred between present-day JMMC and the Faroe Islands (e.g. Gernigon *et al.*, 2015; Blischke *et al.*, 2016).

Our model predicts that, following a major change in extension direction that was coeval with the abandonment of the Labrador Sea-Baffin Bay oceanic spreading and transform system, oblique deformation occurred south of the proto-JMMC and along the poorly defined IFAZ (Fig. 6c). This caused further westward relocation of the spreading centre towards a fossil subduction zone where eclogite and, especially, weak inherited serpentinite accommodated the relocation and final development of the Kolbeinsey Ridge. Complete development of the Kolbeinsey Ridge resulted in final separation of the proto-JMMC from East Greenland (Fig. 6d) and complete breakup of the North Atlantic.

Formation of the JMMC correlates with and can be explained by rejuvenation of pre-existing structures of different ages. Oblique accommodation/deformation zones including fracture zones defined the extent of the JMMC along the spreading axis. This model provides a simple explanation for microplate-complex formation involving control by both plate tectonic processes and structural inheritance.
Further work and data acquisition is required to fully understand the nature and formation of the JMMC, Iceland and the Iceland-Faroe Ridge. All three components are intrinsically interlinked and essential for understanding the tectonic and magmatic evolution of the entire North Atlantic. Geophysical data are lacking especially in the south of the JMMC, offshore northwest Iceland, and between Iceland and the Faroe Islands. The most fundamental and perhaps economically important question is the extent of continental crust underlying this region, a question that may require additional marine surveys, re-interpretation of geochemical data and further drilling and sampling in this area.

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Figures

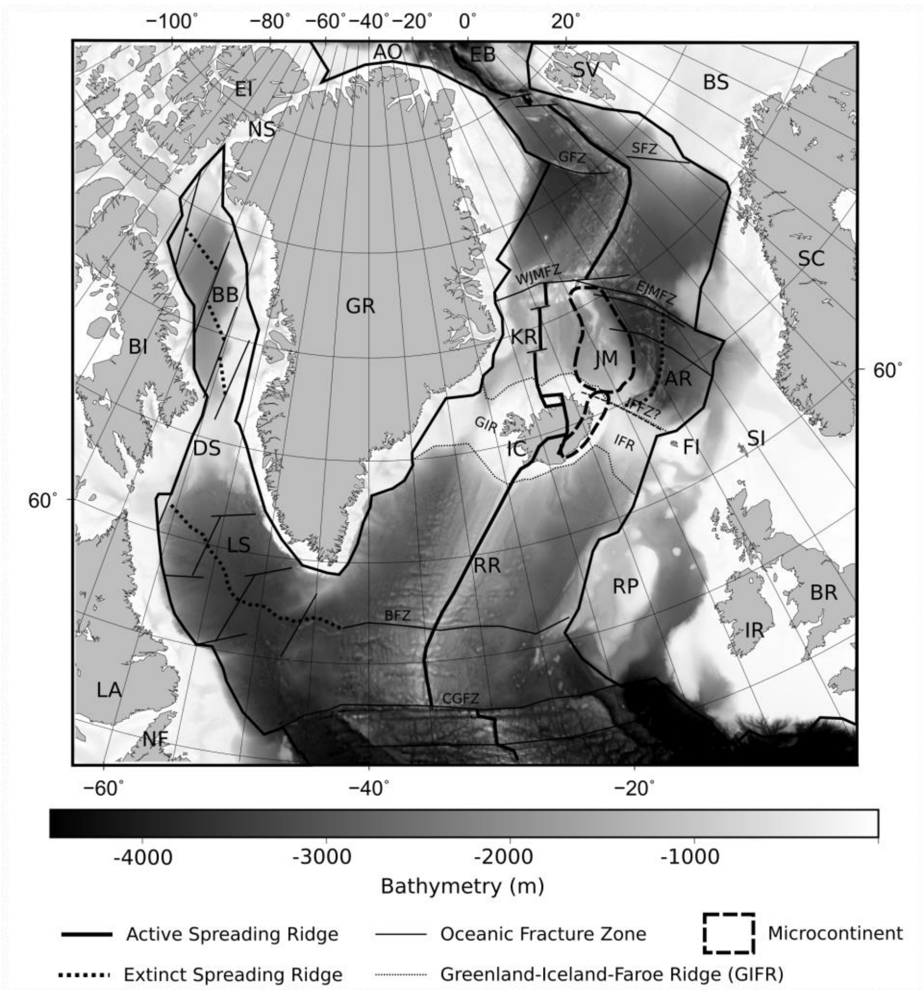

Figure 1

Bathymetric map of the present-day North Atlantic. Bathymetry from the General Bathymetric Chart of the Oceans (GEBCO). Major oceanic fracture zones after Dore *et al.* (2008), Mid Ocean Ridges from Seton *et al.* (2012), microcontinents from Torsvik *et al.* (2015). Greenland-Iceland-Faroe Ridge (GIFR) consists of the Greenland-Iceland Ridge, the Iceland Plateau and the Iceland-Faroe Ridge. The position of the Iceland Faroe Fracture Zone is stippled, but its existence and nature is debated (see text). AO = Arctic Ocean; AR = Aegir Ridge; BB = Baffin Bay; BFZ = Bight Fracture Zone; BI = Baffin Island; BR = Britain; BS = Barents Sea; CGFZ = Charlie-Gibbs Fracture Zone; DS = Davis Strait; EB = Eurasia basin; EI = Ellesmere Island; EJMFZ = East Jan Mayen Fracture Zone; GIR = Greenland-Iceland Ridge; GR = Greenland; IC – Iceland; IFFZ = Iceland-Faroe Fracture Zone; IFR = Iceland-Faroe Ridge; IR = Ireland; KR = Kolbeinsey Ridge; LA = Labrador; LS = Labrador Sea; NF = Newfoundland; NS = Nares Strait; RP = Rockall Plateau; RR = Reykjanes Ridge; SC = Scandinavia; SFZ =

Senja Fracture Zone: SF = Svecofennian; SI = Shetland Islands; SV = Svalbard; WJMFZ = West Jan Mayen Fracture Zone.

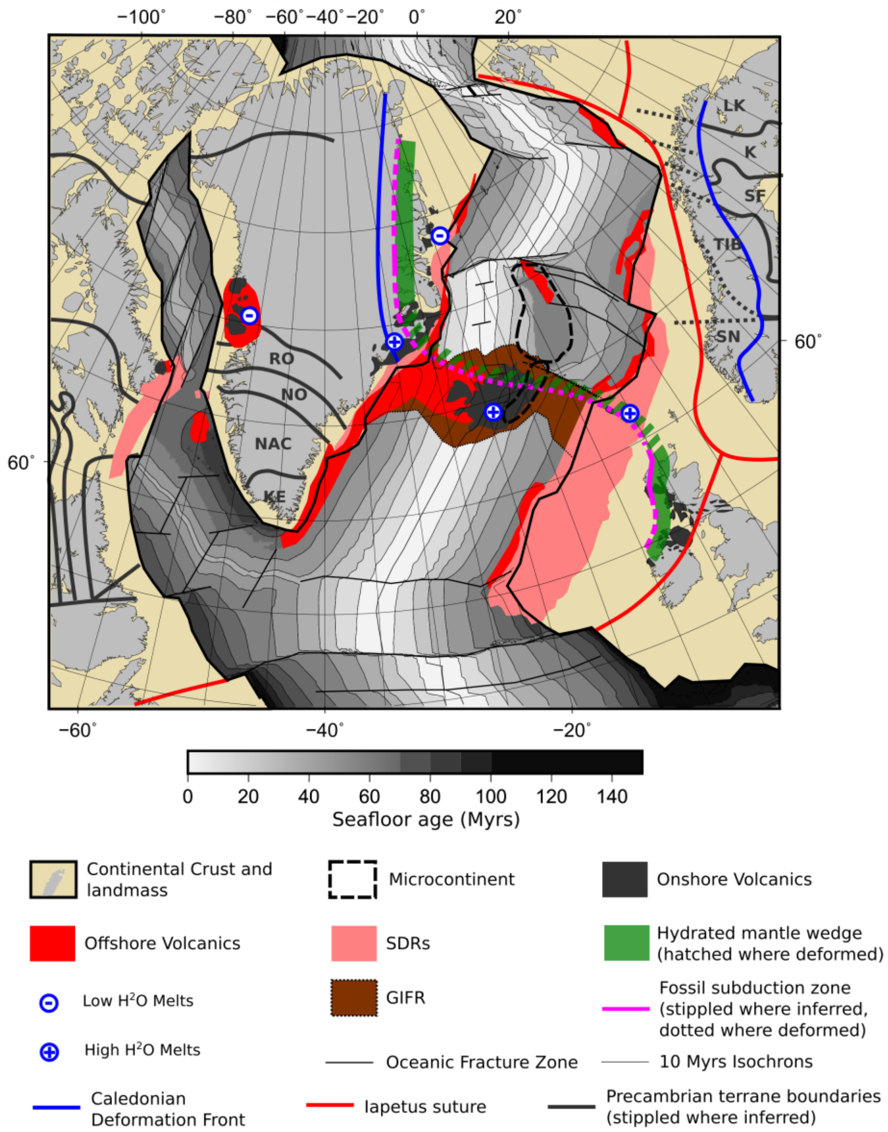

Figure 2

Overview map of the present-day North Atlantic. Seafloor age from Seton *et al.* (2012), major oceanic fracture zones after Doré *et al.* (2008), distribution of igneous rocks of the North Atlantic Igneous Province after Upton (1988), Larsen & Saunders (1998), Abdelmalak *et al.* (2012), Precambrian basement terranes after Balling (2000) and Indrevær *et al.* (2013) – Scandinavia, St-Onge *et al.* (2009) – Greenland and northeastern Canada. Caledonian Deformation Front after Skogseid *et al.* (2000) and

Gee *et al.* (2008). K = Karelian; KE = Ketilian Orogen; LK = Lapland-Kola; NAC = North Atlantic Craton; NO = Nagssugtoqidian Orogen; RO = Rinkian Orogen; SF = Svecofennian; SN = Sveconorwegian: TIB = Transscandinavian Igneous Belt.

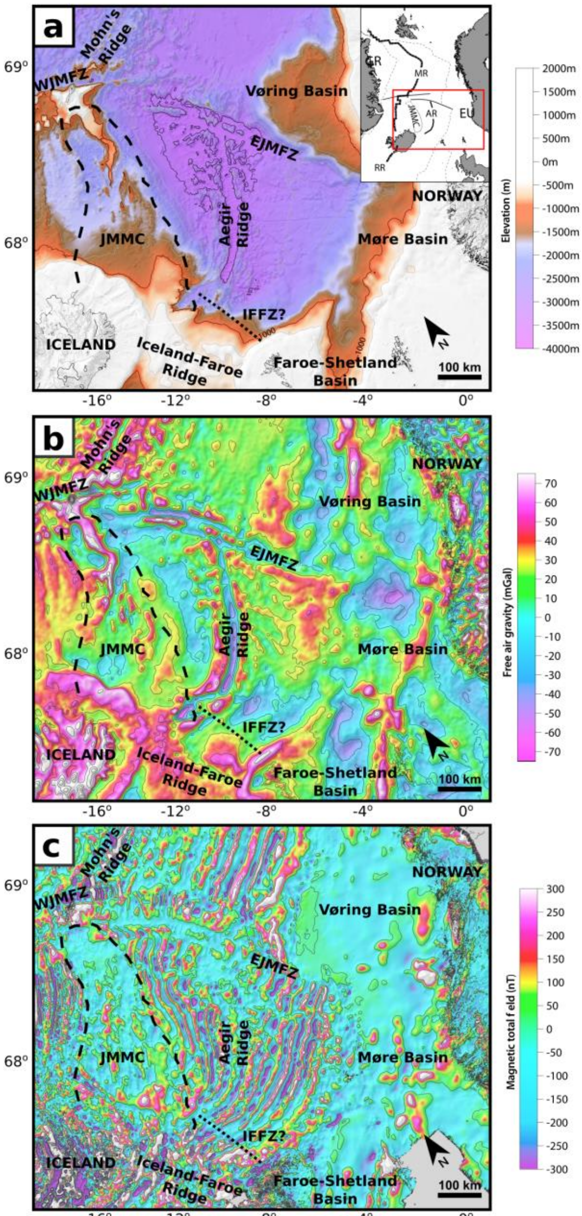

Figure 3

Bathymetry (a), free air gravity (b) and magnetic anomaly (c) maps of the Norway Basin, the Jan Mayen microplate complex (JMMC), Iceland, the Iceland-Faroe Ridge and surrounding conjugate margins (modified after Gernigon *et al.* 2015). The bathymetric map illustrates the special physiological nature of the JMMC, coinciding with large free air gravity anomalies. Magnetic anomalies within the boundaries of the JMMC are weak. This is in large contrast to the adjacent Norway Basin, which shows clear magnetic spreading anomalies, and gravity and topographic anomalies that evidence the “fan-shaped” spreading along the extinct Aegir Ridge. There are vague indications in bathymetry, gravity and magnetic data for the existence of a lineament stretching from the south of the JMMC to the Faroe-Shetland Basin, possibly the IFFZ (Blischke *et al.*, 2016), but the data does not provide indisputable evidence for the existence and the nature of such.

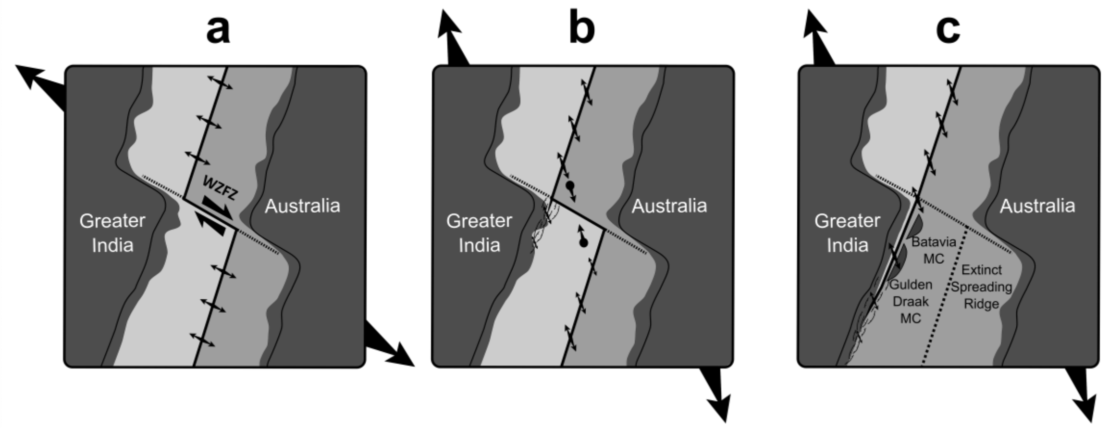

Figure 4

Model for the formation of the Batavia and Gulden Draak microcontinents in the Indian Ocean proposed by Whittaker *et al.* (2016). Initial seafloor spreading occurred perpendicular to the regional plate motions, including the Wallaby-Zenith Fracture Zone (WZFZ). A reconfiguration of plate motions oblique to the developed spreading axes locked the fracture zone, which forced the southern spreading axis to relocate onto a new axis. The new spreading isolates continental fragments (microcontinents) and seafloor spreading separates these from the Indian plate. Large arrows indicate plate motions. Arrows along spreading ridges indicate the spreading direction. Dots with arrows indicate the transpressional regime along the former fracture zone.

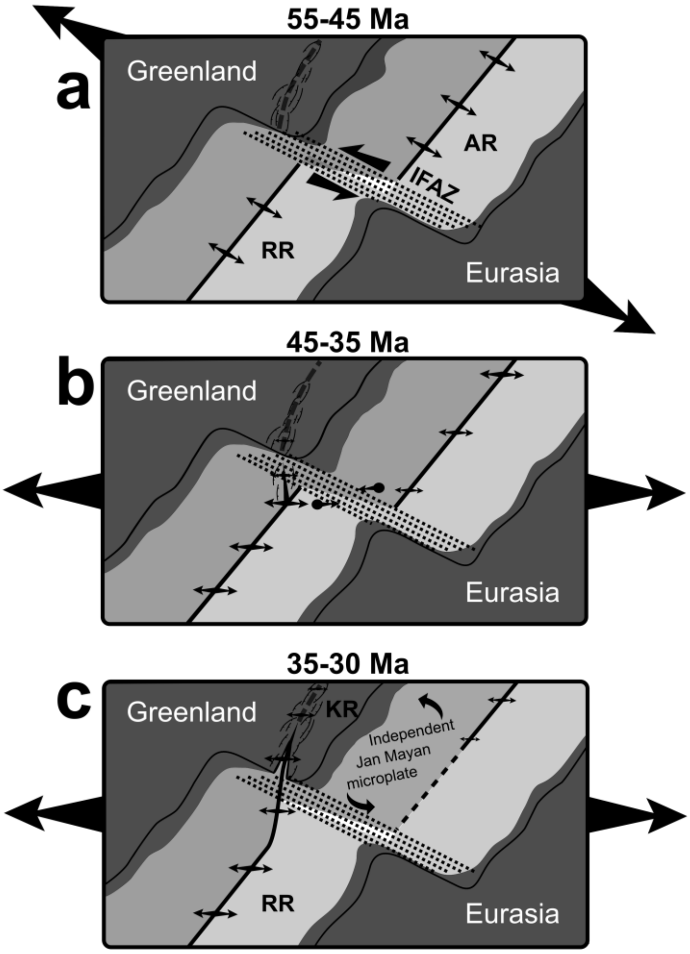

Figure 5

Application of the model of Whittaker *et al.* (2016) to the formation of the Jan Mayen microplate complex. The original model was developed to explain microcontinent separation between Greater India and Australia. (a) NW-SE plate motion between Greenland and Europe with the Iceland-Faroe accommodation zone (IFAZ) as a diffuse zone accommodating relative motion between the Reykjanes ridge (RR) and Aegir ridge (AR). Continental rifting and extension occurs along the lithospheric weakness (East Greenland fossil subduction zone) (b) Plate tectonic reorganisations result in W-E motion between Greenland and Europe locking up the Iceland-Faroe accommodation zone. The Reykjanes ridge diverts towards the north following the lithospheric weakness. (c) Seafloor spreading develops along the Kolbeinsey ridge (KR) breaking the Jan Mayen Microplate off from Greenland. The JMMC rotates counterclockwise. Seafloor spreading on the Aegir ridge is abandoned.

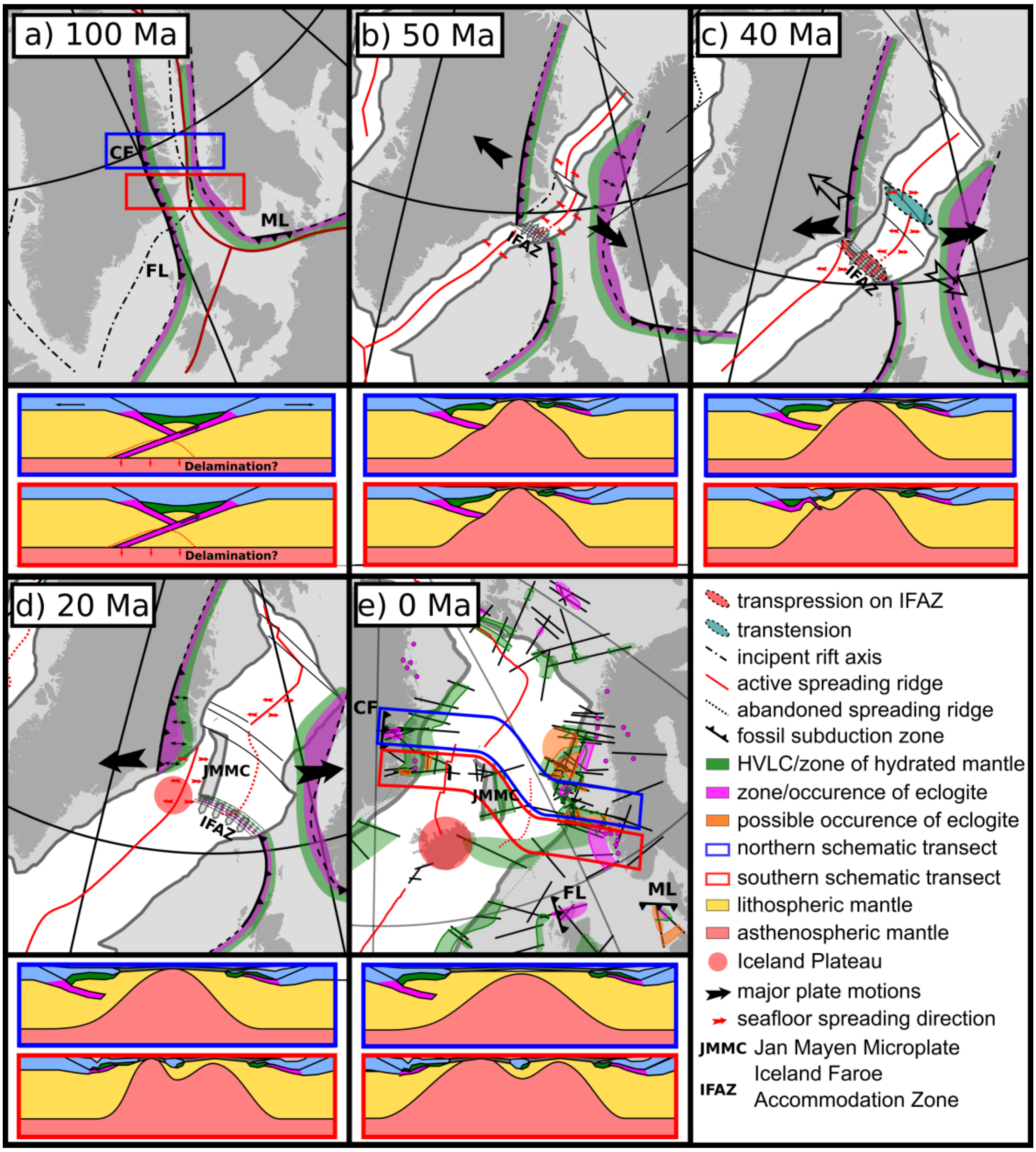

Figure 6:

Separation of the Jan Mayen microplate complex from Greenland. Palaeogeographic reconstructions from Seton *et al.* (2012). 100 Ma: The Caledonian Orogen experienced extensional collapse and multiple rift phases. Fossil subduction zones are still preserved, though possibly deformed. 50 Ma: Seafloor spreading in the North Atlantic separates Greenland from Europe with NW-SE plate motions. Breakup in the NE Atlantic occurs along the Iapetus suture, which deforms. 40 Ma: Plate motions change from NW-SE to W-E, which causes transpression on the Iceland-Faroe accommodation zone. The

Reykjanes ridge spreading centre develops towards the north, following lithospheric weaknesses along the East Greenland fossil subduction zone. 20 Ma: The newly formed Kolbeinsey ridge is almost entirely developed, separating the Jan Mayen Microplate Complex from Greenland. The fossil subduction zone in Central East Greenland is highly deformed, whereas it is mainly preserved further north. The Aegir Ridge is successively abandoned. 0 Ma: Fossil subduction zones are still preserved in East Greenland, northern Scotland and the Danish North Sea sector (Central Fjord structure - CF, Flannan reflector - FL, Mona Lisa structure - ML). In Norway and south-central East Greenland the fossil subduction zone has been destroyed and deformed. It now forms high-seismic-velocity lower crustal bodies that are possible eclogite HVLCBs mapped in magenta and orange).